1   Postseismic gravity change after the 2006-2007 great earthquake doublet and constraints on the

2   asthenosphere structure in the central Kuril Islands

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17   **Abstract**

18   Large earthquakes often trigger viscoelastic adjustment for years to decades depending on the

19   rheological properties and the nature and spatial extent of coseismic stress. The 2006 M_w8.3

20   thrust and 2007 M_w8.1 normal fault earthquakes of the central Kuril Islands resulted in

21   significant postseismic gravity change in GRACE but without a discernible coseismic gravity

22   change. The gravity increase of ~4 μGal, observed consistently from various GRACE solutions

23   around the epicentral area during 2007–2015, is interpreted as resulting from gradual seafloor

24   uplift by ~6 cm produced by postseismic relaxation. The GRACE data are best fit with a model

25   of 25–35 km for the elastic thickness and ~10^{18} Pa s for the Maxwell viscosity of the

26   asthenosphere.   The large measurable postseismic gravity change (greater than coseismic

27   change) emphasizes the importance of viscoelastic relaxation in understanding tectonic

28   deformation and fault-locking scenarios in the Kuril subduction zone.

29

30   **1. Introduction**

31   On 15 November 2006, the shallow part of the central Kuril Islands megathrust (depths <25 km)

32   failed in a great M_w 8.3 earthquake following a swarm of thrust-faulting foreshocks in late

33   September 2006 [*Ammon et al.,* 2008; *Steblov et al,* 2008; *Lay et al.,* 2009]. Following this event,

34   distributed thrust-faulting aftershocks occurred along the forearc; additionally, a parallel band of

35   primarily normal faulting events occurred in the outer rise region. Based on the predicted

36   evolution of stress redistribution following the 2006 thrust event, *Ogata and Toda* [2010]

37   hypothesized that continuous triggered slip occurred on the future fault plane of the January 2007

38   event. On 13 January 2007, the great M_w 8.1 normal faulting earthquake ruptured a steeply

39   dipping fault (depths of 4-33 km) beneath the trench slope [*Lay et al.*, 2009]. Various studies of

40   seismic, geodetic, and tsunami data analysis report moment release between 1.6–5.3×10^{21} N m

41   for the 2006 event and 1.4–2.8×10^{21} N m for the 2007 event [*Steblov et al.,* 2008; *Lay et al.,*

42   2009].

43   Earthquakes with moment magnitudes of $M_w8.1$–8.3 are nearly at, or below, the detection

44   threshold by Gravity Recovery And Climate Experiment (GRACE). Furthermore, the coseismic

45   gravity changes were dominantly a negative anomaly (due to crustal dilatation) following the

46   2006 thrust rupture that was then disrupted by a positive anomaly (crustal compression) after the

47   2007 normal faulting. We calculated a predicted total coseismic gravity change of <1 μGal at

48   GRACE's spatial resolution. This was certainly below the detection threshold [*e.g*., *Han et al*.,

49   2013]. However, following these events, the GRACE Release-05 Level-2 (L2) solutions

50   [*Bettadpur et al*., 2014] exhibited an increasing gravity trend localized around the epicenters of

51   the 2006 and 2007 earthquakes. This positive trend is consistently observed in various other

52   GRACE solutions processed by different groups (Supporting information).   It can be

53   conveniently illustrated by using the CNES/GRGS web tools (http://www.thegraceplotter.com).

54   Compared to the postseismic gravity change after the 2004 Sumatra-Andaman, 2011

55   Tohoku-Oki and 2012 Wharton Basin earthquakes [*e.g*., *Han et al*., 2008, 2014, 2015], gravity

56   change associated with the 2006-2007 Kuril events is distinctly different; namely, the magnitude

57   of postseismic gravity change becomes by far larger than the coseismic change within a few

58   years. We seek to explain such observations by testing alternate viscoelastic models of the

59   asthenosphere beneath an elastic lithosphere. In this study, we address (*i*) how the 2006 $M_w8.3$

60   thrust and 2007 $M_w8.1$ normal faulting earthquakes interplay to produce the observed gravity

61   change, (*ii*) how the rheology of the asthenosphere plays a role in producing postseismic gravity

62   changes, and (*iii*) how GRACE constrains the rheological structure in the central Kuril Islands.

63   We will present the details of GRACE observations and viscoelastic modeling results with a

64   range of rheological parameters (thickness of lithosphere and asthenosphere viscosity) and of

65   coseismic moment magnitude ratios, and compare with the previous Kuril GPS studies of *Kogan*

66   *et al*. [2011, 2013].

67

68   **2. GRACE observation of postseismic gravity change**

69   We used monthly GRACE gravity data from CSR L2 solutions presenting an effective resolution

70   of 500 km (maximum spherical harmonic degree of 40). Although the GRACE RL05 L2 data

71   are provided up to the maximum degree of 60, the resolution to ~300 km is applicable only at the

72   polar region. The monthly gridded data, computed from the truncated spherical harmonic

73   synthesis, were analyzed by applying the least squares fit of the parameters including annual and

74   semi-annual sinusoids, bias, and a Heaviside step and linear trend after year 2007. Figure 1a

75   presents the linear trend estimated between January 2007 and April 2015. The coseismic and

76   postseismic effects after 2011 Tohoku-Oki earthquake dominate the trend map (Figure 1a).

77   However, a statistically significant positive anomaly of 0.5 µGal/year is found around the

78   epicenters of the Kuril earthquakes.

79   We examined the finite fault models by *Ji* [2006, 2007] with the southeast dipping plane

80   for the 2007 normal fault to compute the gravity effect as shown in Figures 1b-d. The synthetic

81   calculation of coseismic (elastic) deformation after the 2006 earthquake results in the negative

82   gravity change northwest of the trench due to crustal dilatation while a positive gravity (due to

83   the hanging wall uplift) is compensated for by the ocean [*e.g.*, *Broerse et al*., 2011] (Figure 1b).

84   This is a typical pattern in coseismic gravity change after thrust earthquakes [*Han et al*., 2013].

85   *Broerse et al*. [2014] found the horizontal motion of the bathymetry is important to understand

86   the cosesimic gravity change. However, the horizontal postseismic motions during the

87   observation period are ~2 orders of magnitude less than the coseismic horizontal motions, and

88   hence they contribute a much smaller fraction of the net gravity change. The 2007 normal fault

89   yielded instead positive gravity due to compression pushed by the downward motion of the

90   hanging wall (Figure 1c). Again, the gravity change due to density change is greater than the

91   gravity change associated with the elastic vertical motion at this spatial scale. The gravimetric

92   signals produced by the two earthquakes compensate each other and resulted in a gravity

93   anomaly smaller than each of the gravimetric signals in the individual earthquakes (Figure 1d).

94   The elastic gravity change is within ±0.5 µGal and appears away from the epicenters at the

95   spatial scale of 500 km. The solutions by *Lay et al*. [2009] including the southeast and northwest

96   dipping planes for the 2007 rupture produced slightly different synthetics, but the above

97   conclusion does not change (Figure S1). Elastic deformation alone does not replicate the peaked

98   positive trend near the epicenter in the GRACE data.

99   The time-series of the gravity change around the peak GRACE signal in Figure 1a

100   (averaged within a spherical cap with a radius of 2.5° centered at 46.5°N, 154.5°E) is shown in

101   Figure 2. The GRACE data (Figure 2, black dotted) presents scatter of 1.8 µGal (root mean

102   square) with respect to the fit (green solid) as described above. The linear trend and annual

103   sinusoid components are statistically significant while the step function and semi-annual

104   components are not. The trend prior to the 2006 earthquake is not statistically different from

105   zero (based on the t-test). The background gravity variation induced by other geophysical

106   processes such as non-tidal atmosphere (6-hourly ECMWF operational data) and ocean mass

107   (OMCT) [*Dobslaw et al*., 2013] is also shown (magenta solid), which was removed *a priori* from

108   GRACE gravity solutions. The amplitude of seasonal atmospheric and oceanic variations is

109 estimated to be ~1 µGal. The GRACE data presents the un-/mis-modeled seasonal change with a

110 amplitude of ~0.5 µGal (green solid). Ignoring the two-month time difference between the 2006

111 and 2007 ruptures, the linear trend after year 2007 is estimated to be 0.45±0.08 µGal/year. The

112 postseismic gravity trend estimate was consistent with the result based on the other GRACE

113 solutions by CNES/GRGS, processed with a different inversion method and different atmosphere

114 and ocean models (Figure S2) and by other groups (Figures S3 and S4).

115 The gravity increase by ~4 µGal over 8 years from GRACE implies ~6 cm of the seafloor

116 uplift around the area of ~500^2 km^2 near the epicenters (0.7 µGal of gravity change per 1 cm of

117 seafloor uplift - the Bouguer correction with the density contrast between the crust and ocean).

118 *Kogan et al.* [2011 and 2013] presented a postseismic uplift rate up to 1.2 cm/year (KURILNET

119 site MATC) from near-field GPS stations along the volcanic arc, which confirmed the

120 postseismic deformation is caused predominantly by viscoelastic relaxation (rather than afterslip).

121 We henceforth assume that the GRACE gravity changes are shaped primarily by viscoelastic

122 relaxation and ignore the possible contribution of afterslip.

123

124 **3. Viscoelastic relaxation modeling**

125 For the Kuril earthquake sequence, the postseismic gravity change has emerged as the GRACE

126 time series have lengthened even though the measured, combined coseismic gravity change was

127 insignificant. In this Section, we examined in what circumstances the postseismic gravity

128 changes could be larger than the cosesimic gravity change within a few years after the rupture by

129 quantifying the gravity response functions (Green's functions) of the viscoelastic Earth. We

130 delineate the overall range of viscoelastic parameters important to control the postseismic gravity

131 change by using a simplified point earthquake source and testing various synthetic structures.

132   Following *Han et al.* [2013], the gravity change by an earthquake point source is

133   expressed as

$$g(\theta,\phi,t) = \left(\frac{-M_{rr}}{2}\right)g_{Mrr}(\theta,\phi,t) + \left(M_{r\theta}\right)g_{Mr\theta}(\theta,\phi,t) + \left(M_{r\phi}\right)g_{Mr\phi}(\theta,\phi,t)$$

134
$$+ \left(\frac{M_{\theta\theta} - M_{\phi\phi}}{2}\right)g_{M\theta\theta-M\phi\phi}(\theta,\phi,t) + \left(M_{\theta\phi}\right)g_{M\theta\phi}(\theta,\phi,t) \qquad , \qquad (1)$$

135   where a dislocation source is assumed to be at the north pole and described with the moment

136   tensor components, M_{rr}, $M_{r\theta}$, $M_{r\phi}$, $M_{\theta\theta} - M_{\phi\phi}$, and, $M_{\theta\phi}$. The respective Green's function is

137   g_{Mrr}, $g_{Mr\theta}$, $g_{Mr\phi}$, $g_{M\theta\theta-M\phi\phi}$, and $g_{M\theta\phi}$. They are dependent on the Earth's viscoelastic structure

138   and the depth of dislocation source. The first one g_{Mrr} is constructed with the $m=0$ (monopole

139   source) eigenfunctions of viscoelastic normal modes [*Pollitz*, 1997; *Pollitz et al.*, 2006]. The

140   pair of $g_{Mr\theta}$ and $g_{Mr\phi}$ is determined by the $m=1$ (dipole source) eigenfunctions. The other pair

141    $g_{M\theta\theta-M\phi\phi}$ and $g_{M\theta\phi}$ is computed by the $m=2$ (quadrupole source) eigenfunctions. Therefore, there

142   are in principle only three independent functions that govern temporal changes of gravity and

143   displacement fields for each source depth with a given viscoelastic structure.

144   Using the normal mode approach, we computed the viscoelastic (postseismic) and elastic

145   (coseismic) Green's functions. The seismic source was assumed to be at 20 km depth within the

146   elastic layer. The viscosity for the upper mantle (220–670 km) and for the lower mantle (670–

147   2891) was fixed to be 10^{20} Pa s and 10^{21} Pa s, respectively. The most sensitive parameters to

148   govern temporal evolution of gravity are the asthenosphere viscosity and the lithosphere (elastic)

149   thickness. Three different elastic thicknesses, 25, 40, and 115 km, were tested; below these

150   depths the viscoelastic asthenosphere extends down to 220 km. Two different Maxwell

151   viscosities, 10^{18} and 10^{19} Pa s, for the asthenosphere were tested (Table 1 of Supporting

152   Information).

153   Figure 3 shows the ratio of the postseismic gravity response 5 years after the rupture

154   relative to the coseismic change at each spherical harmonic degree (l) for the m=0, m=1, and

155   m=2 responses, computed from six different cases for the viscoelastic structure. Note that such

156   ratio is not sensitive to the seismic source parameters (except depth), but only to the Earth

157   structure. Generally speaking, the shallower asthenosphere and the lower viscosity yield larger

158   changes in the ratios. The ratio of the m=1 response is significantly smaller than the other

159   response functions for all cases, which implies that the viscoelastic relaxation is governed more

160   effectively by m=0 and m=2 patterns. We found that the asthenosphere viscosity of 10^{19} Pa s

161   does not produce the postseismic gravity change that is significantly larger than the coseismic

162   change regardless of the thickness of the overlying elastic layer. However, the asthenosphere

163   viscosity of 10^{18} Pa s and an elastic layer thickness of <40 km can produce an order-of-

164   magnitude larger postseismic gravity change than the coseismic gravity in 5 years. We extend

165   the viscoelastic modeling to constrain such parameters in the following Section.

166

167   **4. Constraints on the asthenosphere structure**

168   We exploit the GRACE gravity observation to deduce the optimal parameters that may represent

169   the rheological structure in the central Kuril Islands region. First, it is necessary to fix the

170   seismic sources of the 2006 and 2007 earthquakes. *Lay et al*. [2009] indicates a large variability

171   in seismic and geodetic source models (point sources as well as finite faults shown in their

172   Tables 1 and 2). The moment release from the 2006 rupture has been estimated to be M_0=1.6–

173   5.3×10^{21} N m including the Global Centroid Moment Tensor (GCMT) M_0=3.5×10^{21} N m [*Lay et*

174   *al*., 2009]. For the 2007 normal fault, it is M_0=1.4–2.8×10^{21} N m with GCMT M_0=1.8×10^{21} N m.

175   The moment release ratio of 2006 to 2007 ruptures varies from 1.1 to 3.8. Considering such

176 large uncertainty in the moment release from the two events, we decided to adjust the moment

177 release ratio in addition to the viscoelastic structure parameters. Due to limited detection

178 threshold of the GRACE measurements, it is not possible to constrain the moment release for

179 each earthquake separately.

180   The synthetic gravity change was computed using the finite fault models of *Ji* [2006,

181 2007] and the software VISCO1D for viscoelastic deformation [*Pollitz*, 1997]. Figure 4 shows

182 the average rate of postseismic gravity change over 2007–2015 for the exemplary case of an

183 elastic thickness of 32 km and the asthenosphere Maxwell viscosity of 10^{18} Pa s; the deeper (>

184 220 km) upper mantle and lower mantle are as described in Section 3. The 2006 fault yields the

185 positive gravity change primarily due to vertical motion of various density interfaces (such as

186 seafloor and Moho) because viscoelastic deformation is shown to be nearly incompressible [*e.g.*,

187 *Han et al.*, 2014, 2015]. The top of the asthenosphere is subject to the boundary condition of

188 broad-scale shortening, which triggers postseismic uplift [*Melosh*, 1983]. In contrast, the 2007

189 normal fault (for both northwest and southeast dipping planes) yields negative gravity change

190 associated with postseismic subsidence under the top boundary of the asthenosphere subject to

191 extension (See the schematic diagrams in Figure 4). The actual 2007 rupture, along with the pre-

192 slip on this normal faulting rupture plane, in the outer rise caused contrasting vertical

193 deformation to that triggered by the 2006 earthquake at the plate interface.

194   We write the total postseismic gravity change, that is comparable to GRACE observation,

195 from a given viscoelastic model as follows:

196    $$g(t) = (1+\alpha) \cdot g_{2006}(d_a, v, t) + (1+\beta) \cdot g_{2007}(d_a, v, t),$$ (2)

197 where the synthetic gravity changes g_{2006} and g_{2007} are evaluated at time t with the viscosity v

198 and the elastic thickness d_a from the 2006 and 2007 ruptures, respectively. The unknown scale

199   factors, α and β, are introduced to accommodate the uncertainty in the moment magnitude in

200   each seismic model, respectively. They are not constrained independently but in terms of a ratio

201    $\gamma = \dfrac{(1+\alpha)}{(1+\beta)} \dfrac{M_0^{2006}}{M_0^{2007}}$, where M_0^{2006} and M_0^{2007} is the scalar moment for the 2006 and 2007

202   earthquake determined from *a priori* seismic models, *e.g.*, *Ji* [2006, 2007], respectively. The

203   ratio implies the relative size of the moment release of the 2006 to the 2007 rupture. We tested

204   these three parameters within a reasonable range; *i.e.*, d_a=26–88 km, v=10^{17}–10^{19} Pa s, and γ

205   =1.0–4.0.

206   We computed numerous cases of synthetic gravity changes and compared with the

207   GRACE data. We quantified the goodness of the fit by computing the R^2 statistics (*i.e.*, variance

208   reduction). Figures 5a-c illustrate the R^2 statistics with variable viscosities and moment ratios,

209   respectively, for d_a = 26, 32, and 38 km. The results of d_a>40 km were ruled out since the fits

210   were significantly degraded within the specified range of v and γ. The best fits are found

211   consistently when the viscosity is close to 10^{18} Pa s. The larger elastic layer thickness d_a

212   requires the larger moment ratio γ (The identical results can be drawn based on the reduced chi-

213   squared χ^2 statistics). The long-period seismic estimates of the relative moment ratio between

214   the 2006 and 2007 ruptures are within 2–3 [*T. Lay*, 2015, personal communication]. With this

215   seismic constraint, the case of d_a=38 km can also be ruled out. The GRACE data are best fit

216   with a model of 25–35 km for the elastic thickness and ~10^{18} Pa s for the Maxwell viscosity of

217   the asthenosphere.

218   The gravity time-series from three best-fit model parameters (color crosses in Figure 5a-

219   c) were compared with the GRACE time series (Figure 5d). All cases reasonably fit the data

220 yielding $R^2 > 0.75$ and $\chi^2 \sim 1.0$. We also tested the biviscous (Burgers) rheology that implements

221 the transient (Kelvin) element in series with the steady-state (Maxwell) element [*e.g.*, *Han et al.*,

222 2014, 2015]. The identical shear modulus was used for the Kelvin and Maxwell elements. We

223 found that the transient viscosity is not as important as the steady-state viscosity to explain the

224 GRACE data in 2007–2015. Any biviscous model with a Maxwell viscosity of 10^{18} Pa s and a

225 Kelvin viscosity of 1–10×10^{17} Pa s fits as well as the Maxwell rheology model with 10^{18} Pa s

226 (Figure 5d). The Burgers or Maxwell model with the steady-state viscosity of 10^{19} Pa s, however,

227 significantly underpredicts the gravity change (<1 μGal in 8 years) and does not agree with the

228 GRACE measurement (~0.5 μGal/year). The spatial pattern of the average rate of the synthetic

229 gravity changes in 2007–2015 is shown in Figure 5e using the model parameters of d_a=32 km,

230 v=10^{18} Pa s, and γ=3.0. The primary positive gravity around the epicenters is consistent with

231 the GRACE data in terms of the spatial location, the anomaly magnitude, and the temporal

232 pattern, while the secondary smaller anomaly of the negative change is less discernable.

233

234 **5. Discussion**

235 Based on the analysis of 8 years of continuous gravity data, we estimated a Maxwell steady-state

236 viscosity of 10^{18} Pa s for the asthenosphere, agreeing with *Kogan et al.* [2013]. Based on their

237 analysis of 4 years postseismic GPS displacement data, *Kogan et al.* [2013] conjectured the

238 Maxwell asthenospheric viscosity is lower than that in Chile and Alaska (10^{19} Pa s), but these

239 events were much larger in terms of spatial extent and moment magnitude. With the additional

240 seismic constraints on the relative moment between two ruptures, our gravity analysis indicates

241 the elastic layer thickness confined to be <35 km, which may be largely influenced by the mantle

242 wedge beneath the active Kuril volcanic arc. Our preferred elastic thickness is smaller than 62

243   km fixed (not adjusted) by *Kogan et al.* [2011, 2013] for their GPS analysis, and rather

244   consistent with the oceanic lithosphere thickness (30–40 km) from the flexural analysis of the

245   bathymetry around the Kuril trench [*Hanks*, 1971] and with the numerical modeling beneath

246   adjacent northeastern Japan [*Muto et al.*, 2013 and *Hu et al.*, 2014].

247   Lateral variations in rheology are likely around subduction zones, as differences in elastic

248   plate thickness, elastic moduli and viscosity structure are expected between the landward and

249   oceanic side of the subducting slab (*e.g.*, *Billen and Gurnis*, [2003]; *Abers et al.*, [2006]). This

250   variation and the presence of the slab can substantially modify predicted postseismic deformation

251   (*e.g.*, *Cohen et al.*, [1996]; *Pollitz et al.*, [2008]; *Hu and Wang*, [2012]) and presumably the

252   gravity signal as well.

253   In our modeling, we neglected the lateral heterogeneity of the Earth's structure that may

254   bias our estimates of the elastic layer thickness. For example, the 2007 outer rise fault occurred

255   in a thicker elastic layer [*Lay et al.*, 2009] and is expected to have produced smaller viscoelastic

256   relaxation than in a thinner elastic layer, which would lead to less disruption of the viscoelastic

257   uplift produced by the 2006 rupture. However, the Maxwell asthenosphere viscosity estimate is

258   consistent regardless of the assigned thickness of the overlying elastic lithosphere. As depicted

259   in Figure 4, the 2006 rupture triggers an initially large viscoelastic uplift while the 2007 rupture

260   adds the extensional boundary condition that pulls down the triggered uplift. What GRACE

261   observed is the remnant of those contrasting postseismic vertical motions.

262   As the gravity time-series lengthen, the sensitivity increases more drastically for

263   detecting the postseismic trend than the coseismic change. Due to relocking of the shallow

264   segment of the plate interface, a small part (as yet not discernible) of the postseismic gravity

265   change will be constant interseismic deformation as well. While the magnitude of coseismic

266   gravity change is proportional to moment magnitude, postseismic (viscoelastic) gravity change is

267   also dependent on the rheological structures that are variable from region to region. Our analysis

268   of the M_w8.3 2006 and M_w8.1 2007 Kuril earthquakes prompts reexamination of the GRACE

269   data particularly with regard to postseismic gravity changes associated with even smaller

270   earthquakes including the M_w8.1 2009 Samoa-Tonga earthquake doublet. This will become

271   increasingly relevant as the GRACE time-series lengthens and the time series of gravity is

272   extended with GRACE Follow-On starting in 2017 [*Flechtner, et al*., 2014].

273   This 2006-2007 Kuril sequence is one of five earthquake sequences that has produced

274   persistent viscoelastic relaxation signals in GRACE gravity; the other cases are those following

275   the 2004, 2010, 2011, and 2012 great earthquakes [*e.g.*, *Han et al*., 2014, 2015, *Broerse et al*.,

276   2015]. The prevalence of viscoelastic relaxation was also observed in the seafloor data above the

277   2011 Tohoku-Oki rupture [*Sun et al*., 2014]. The continuing analysis of the Earth's rheological

278   structure is imperative to understanding of the complete earthquake cycle processes. *Wang et al*.

279   [2013] emphasized the importance of viscoelastic relaxation in drawing an adequate picture of

280   fault-locking distributions at megathrust boundaries, noting that incorrect assessment of seismic

281   and tsunami hazards may result from not considering the Earth's rheological response. In

282   addition, understanding of the Earth's rheological response is also crucial for estimating climate

283   signals (water and ice mass change) as the elastic and viscoelastic (un-)loading effects that exist

284   in geodetic data such as GPS and GRACE need to be removed [*Ivins et al.*, 2011].

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References

Abers, G.A., van Keken, P.E., Kneller, E.A., Ferris, A., Stachnik, J.C. (2006), The thermal structure of subduction zones constrained by seismic imaging: implications for slab dehydration and wedge flow, *Earth Planet Sci. Lett.*, 241, 387–397.

Ammon, C. J., Kanamori, H., & Lay, T. (2008). A great earthquake doublet and seismic stress transfer cycle in the central Kuril islands. *Nature*, *451*(7178), 561-565.

Bettadpur et al. (2014), CSR Level-2 Status, a paper presented at GRACE Science Team Meeting, Potsdam, Germany, September 29-31, 2014

Billen,M.I., Gurnis,M. (2003), Comparison of dynamic flow models for the Central Aleutian and Tonga–Kermadec subduction zones. *Geochem. Geophys. Geosyst.* 4 (4), 1035. doi:10.1029/2001GC000295.

Broerse, D., L. Vermeersen, R. Riva, and W. van der Wal (2011), Ocean contribution to co-seismic crustal deformation and geoid anomalies: Application to the 2004 December 26 Sumatra-Andaman earthquake, *Earth Planet. Sci. Lett.*, 305(3–4), 341–349, doi:10.1016/j.epsl.2011.03.011.

Broerse, T., R. Riva, and B. Vermeersen (2014), Ocean contribution to seismic gravity changes: The sea level equation for seismic perturba- tions revisited, *Geophys. J. Int.*, *199*(2), 1094–1109.

Broerse, T., R. Riva, W. Simons,□R. Govers, and B. Vermeersen□(2015), Postseismic GRACE and GPS observations indicate a rheology contrast above and below the Sumatra slab, J. Geophys. Res. Solid Earth, 120, doi:10.1002/2015JB011951. □

316   Cohen, S.C. (1996), Time-dependent uplift of the Kenai Peninsula and adjacent regions of south

317   central Alaska since the 1964 Prince William Sound earthquake, J. geophys. Res., 101,

318   8595–8604.

319   Dobslaw, H., Flechtner, F., Bergmann-Wolf, I., Dahle, Ch., Dill, R., Esselborn, S., Sasgen, I.,

320   Thomas, M. (2013), Simulating High-Frequency Atmosphere-Ocean Mass Variability for

321   De- Aliasing of Satellite Gravity Observations: AOD1B RL05, J. Geophys. Res., 118(C5),

322   10.1002/jgrc.20271.

323   Flechtner, F., Morton, P., Watkins, M., & Webb, F. (2014). Status of the GRACE follow-on

324   mission. In: Gravity, Geoid and Height Systems (pp. 117-121). Springer International

325   Publishing.

326   Han, S.-C., J. Sauber, S. B. Luthcke, C. Ji, and F. F. Pollitz (2008), Implications of postseismic

327   gravity change following the great 2004 Sumatra-Andaman earthquake from the regional

328   harmonic analysis of GRACE intersatellite tracking data, J. Geophys. Res., 113, B11413,

329   doi:10.1029/2008JB005705.

330   Han, S.-C., R. Riva, J. Sauber, and E. Okal (2013), Source parameter inversion for recent great

331   earthquakes from a decade-long observation of global gravity fields, J. Geophys. Res.

332   Solid Earth, 118, 1240–1267, doi:10.1002/jgrb.50116.

333   Han, S.-C., J. Sauber, and F. Pollitz (2014), Broadscale postseismic gravity change following the

334   2011 Tohoku-Oki earthquake and implication for deformation by viscoelastic relaxation

335   and afterslip, Geophys. Res. Lett., 41, 5797–5805, doi:10.1002/2014GL060905.

336   Han, S.-C., J. Sauber, and F. Pollitz (2015), Coseismic compression/dilatation and viscoelastic

337   uplift/subsidence following the 2012 Indian Ocean earthquakes quantified from satellite

338   gravity observations, Geophys. Res. Lett., 42, doi:10.1002/2015GL063819.

339   Hanks, T. C. (1971), The Kuril Trench - Hokkaido Rise System: Large Shallow Earthquakes and

340   Simple Models of Deformation, Geophys. J. R. astr. Soc., 23, 173-189.

341   Hu, Y., and K. Wang (2012), Spherical-Earth finite element model of short-term postseismic

342   deformation following the 2004 Sumatra earthquake, J. Geophys. Res., 117(B5), B05404,

343   doi:10.1029/ 2012JB009153.

344   Hu, Y., R. Burgmann, J. Freymueller, P. Banerjee, and K. Wang (2014), Contributions of

345   poroelastic rebound and a weak volcanic arc to the postseismic deformation of the 2011

346   Tohoku earthquake, Earth, Planets and Space, 66:106.

347   Ivins, E., et al. (2011), Onland ice loss and glacial isostatic adjustment at the Drake Passage, J.

348   Geophys. Res, 116, B02403.

349   Ji, C., (2006), Rupture process of the 2006 Nov 15 magnitude 8.3 - Kuril Island earthquake,

350   available at http://earthquake.usgs.gov/eqcenter/eqinthenews/2006/usvcam/finite_fault.php

351   Ji, C., (2007), Rupture process of the 2007 Jan 13□magnitude 8.1 - Kuril Island earthquake, at

352   http://earthquake.usgs.gov/eqcenter/eqinthenews/2007/us2007xmae/finite_fault.php

353   Kogan, M. G., N. F. Vasilenko, D. I. Frolov, J. T. Freymueller, G. M. Steblov, B. W. Levin, and

354   A. S. Prytkov (2011), The mechanism of postseismic deformation triggered by the 2006–

355   2007 great Kuril earthquakes, Geophys. Res. Lett., 38, L06304,

356   doi:10.1029/2011GL046855.

357   Kogan, M. G., N. F. Vasilenko, D. I. Frolov, J. T. Freymueller, G. M. Steblov, A. S. Prytkov,

358   and G. Ekstrom (2013), Rapid postseismic relaxation after the great 2006–2007 Kuril

359   earthquakes from GPS observations in 2007–2011, J. Geophys. Res. Solid Earth, 118,

360   3691–3706, doi:10.1002/jgrb.50245.

361 Lay, T., Kanamori, H., Ammon, C. J., Hutko, A. R., Furlong, K., & Rivera, L. (2009). The

362   2006–2007 Kuril Islands great earthquake sequence. Journal of Geophysical Research:

363   Solid Earth (1978–2012), 114(B11).

364 Melosh, H. J. (1983), Vertical movements following a dip-slip earthquake, Geophys. Res. Lett.,

365   10, 47–50, doi:10.1029/GL010i001p00047.

366 Muto, J., B. Shibazaki, Y. Ito, T. Iinuma, M. Ohzono, T. Matsumoto, and T. Okada (2013), Two-

367   dimensional viscosity structure of the northeastern Japan islands arc-trench system,

368   Geophys. Res. Lett., 40, 4604–4608, doi:10.1002/grl.50906.

369 Ogata, Y., & Toda, S. (2010). Bridging great earthquake doublets through silent slip: On- and

370   off-fault aftershocks of the 2006 Kuril Island subduction earthquake toggled by a slow slip

371   on the outer rise normal fault of the 2007 great earthquake. Journal of Geophysical

372   Research: Solid Earth (1978–2012), 115(B6).

373 Panet, I., Pollitz, F.F., Mikhailov, V., Diament, M., Banerjee, P., and Grijalva, K. (2010), Upper

374   mantle rheology from GRACE and GPS postseismic deformation after the 2004 Sumatra-

375   Andaman earthquake, Geochem. Geophys. Geosyst., Q06008, doi:10.1029/2009GC002905.

376 Pollitz, F. F. (1997), Gravitational-viscoelestic postseismic relaxation on a layered spherical

377   Earth, J. Geophys. Res., 102, 17,921–17,941, doi:10.1029/97JB01277.□

378 Pollitz, F., R. Bürgmann, and P. Banerjee (2006), Postseismic relaxation following the great

379   2004 Sumatra-Andaman earthquake on a compressible self-gravitating Earth, Geophys. J.

380   Int., 167, 397–420.□

381 Pollitz, F., P. Banerjee, K. Grijalva, B. Nagarajan, and R. Bürgmann and (2006), Effect of 3-D

382   viscoelastic structure on post-seismic relaxation from the 2004 M = 9.2 Sumatra

383   earthquake, *Geophys. J. Int*., 173, 189–204.

384   Steblov, G. M., Kogan, M. G., Levin, B. V., Vasilenko, N. F., Prytkov, A. S., & Frolov, D. I.

385   (2008). Spatially linked asperities of the 2006–2007 great Kuril earthquakes revealed by

386   GPS. Geophysical Research Letters, 35(22).

387   Sun, T., et al. (2014), Prevalence of viscoelastic relaxation after the 2011 Tohoku-oki earthquake,

388   Nature, 514, 84–87.

389   Wang, K., Y. Hu, and J. He (2013), Deformation cycles of subduction earthquakes in a

390   viscoelastic Earth, Nature, 484, 327–332.

391

392   **Figure 1**. (a) The spatial pattern of postseismic gravity change from the linear trend of GRACE

393   gravity data between 2007 and 2015 [μGal/year]. The sizeable 2011 Tohoku-Oki earthquake

394   effect can be seen, southwest of the Kuril region. In this example, the GRACE solutions from

395   CSR were used up to degree and order 40 equivalent to 500 km in spatial resolution; (b) and (c)

396   The synthetic coseismic gravity change [μGal] after the 2006 thrust rupture and 2007 normal

397   fault, respectively, computed using the seismic finite fault models and the isotropic PREM

398   elastic structure; (d) The combined coseismic gravity change [μGal]. The calculated coseismic

399   (elastic) gravity change alone does not account for the observed overall GRACE signals.
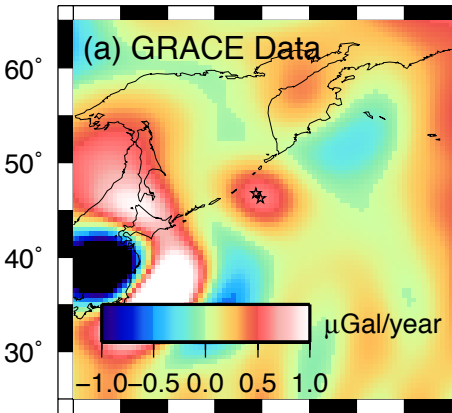
400

401   **Figure 2.** The time series of the GRACE data around the peak GRACE signal in Figure 1a

402   (within a spherical cap with a radius of 2.5° centered at 46.5°N, 154.5°E). The coseismic

403   correction of the 2011 Tohoku-Oki earthquake (-0.5 μGal) was applied in the GRACE time-

404   series following *Han et al*. [2014]. The GRACE data is shown in black with its error estimate,

405   the data fit in green, and the *a priori* model variation of non-tidal atmosphere and ocean mass

406   variability in magenta. The postseismic trends are by far larger than the background atmosphere

407   and ocean mass variability.

408

409   **Figure 3.** The ratio of the Green's (eigen) functions between postseismic (viscoelastic) and

410   coseismic (elastic) gravity change in 5 years after the rupture, respectively, in (a), (b), and (c),

411   for the response functions of the monopole (m=0), dipole (m=1), and quadrupole (m=2) sources.

412   The ratios are depicted for different spherical harmonic degrees up to 50. An earthquake point

413   source is assumed at depth 20 km. For computation of viscoelastic normal mode relaxation,

414   three different thicknesses of the elastic layer and two different viscosities of the underlying

415   asthenosphere were tested. In the monopole and quadrupole gravity responses, the postseismic

416   gravity change was up to one order of magnitude greater than the coseismic gravity at the

417   GRACE spatial scale, depending on the rheological structures.

418

419   **Figure 4.** (a) The synthetic viscoelastic gravity change of the 2006 thrust fault in terms of the

420   linear trend between 2007 and 2015 [μGal/year]; (b) The same but for the 2007 normal fault.

421   The Maxwell asthenosphere viscosity of 10^{18} Pa s and the elastic thickness of 32 km were used

422   in this example. Schematic illustrations of the positive (and negative) gravity rate due to

423   postseismic uplift (and subsidence) under different boundary conditions of the thrust (and

424   normal) faulting imposed on the top of the asthenosphere.

425

426   **Figure 5**. (a)-(c) The R^2 statistics between the GRACE gravity data and various synthetics. The

427   R^2 value of 0.7 means the variance of the GRACE data are reduced by 70% by removing the

428   model time-series from the GRACE data. Three different model parameters were adjusted;

429   elastic thickness d_a, relative moment release γ, and asthenosphere viscosity v; (d) The time-

430   series of the GRACE data, as in Figure 2, and the best-fit Maxwell rheology models (as marked

431   with the color crosses in 5a-c) for the elastic thicknesses of 26, 32, and 38 km. Two examples of

432   the Burgers rheology model with the steady-state viscosity of 10^{18} Pa s and the elastic thickness

433   of 32 km and with the transient viscosity of 10^{17} Pa s (Burgers1) and 10^{18} Pa s (Burgers2) are

434   also shown. The synthetics with the thickness >40 km deviate significantly from the GRACE

435   data and thus were ruled out and are not shown; (e) The spatial pattern of the synthetic

436   viscoelastic gravity change from the best-fit model parameters, as marked with the green cross in

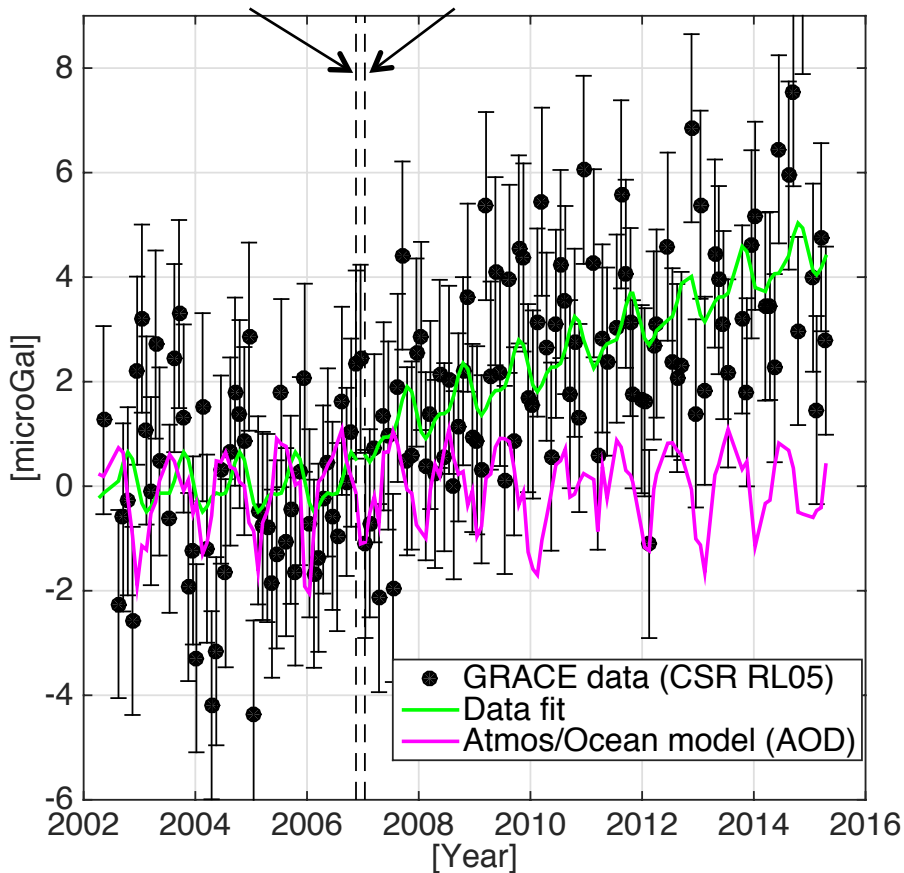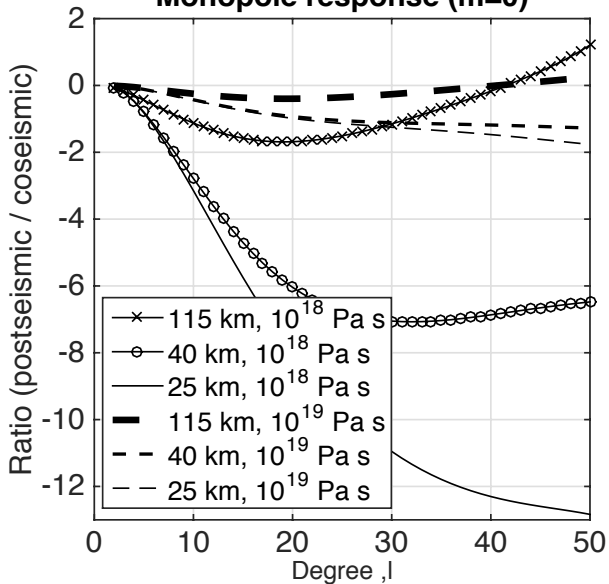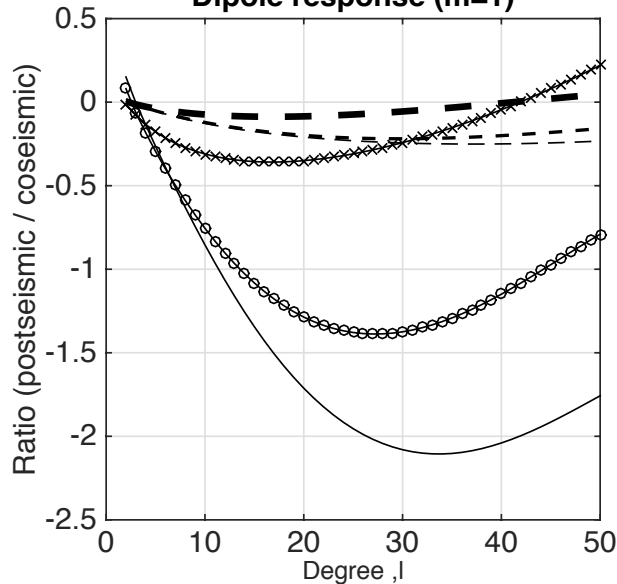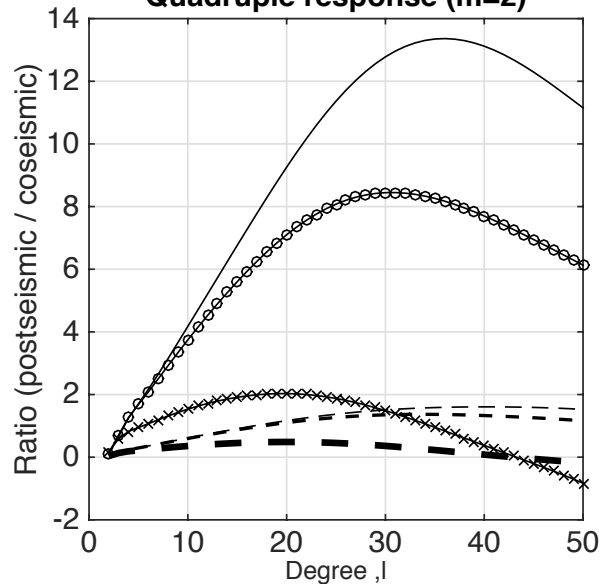437   (b). This is comparable to the GRACE data shown in Figures 1a.

(a) GRACE Data — μGal/year

(b) 2006 Thrust Fault — μGal — Mw 8.3

(c) 2007 Normal Fault — μGal — Mw 8.1

(d) 2006+2007 Faults — μGal — Mw 8.3, Mw 8.1

Monopole response (m=0)

Ratio (postseismic / coseismic) vs Degree ,l

- —×— 115 km, 10^{18} Pa s
- —○— 40 km, 10^{18} Pa s
- —— 25 km, 10^{18} Pa s
- **——** 115 km, 10^{19} Pa s
- **- - -** 40 km, 10^{19} Pa s
- - - - 25 km, 10^{19} Pa s

Dipole response (m=1)

Ratio (postseismic / coseismic) vs Degree ,l

Quadruple response (m=2)

Ratio (postseismic / coseismic) vs Degree ,l

1,000 km   400 km

(a) 2006 Thrust Fault

(b) 2007 Normal Fault

Elastic layer

Boundary condition
Net shortening
=> Uplift
=> Gravity increase

Viscoelastic layer

Elastic layer

or

Boundary condition
Net extension
=> Subsidence
=> Gravity decrease

Viscoelastic layer

(a) d_a = 26 km (b) d_a = 32 km (c) d_a = 38 km

(d) Gravity time-series

GRACE
26 km, Maxwell
32 km, Maxwell
38 km, Maxwell
32 km, Burgers1
32 km, Burgers2

(e) Viscoelastic Gravity

μGal/year